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# OCEANOGRAPHY OF THE EASTERN EQUATORIAL PACIFIC OCEAN*

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According to its circulation the equatorial Pacific Ocean can be divided into three parts. In the central part circulation is predominantly zonal and all currents are well established; in the eastern part this zonal flow pattern is formed by currents flowing toward the equator; in the western part the system of zonal currents disintegrates and water masses flow away from the equator. The transition from the eastern to the central part takes place between about 125° W. and 140° W., and at 140° W. the zonal pattern characteristic for the central equatorial Pacific Ocean is well established. This review will therefore include the area between the coast of America and 140° W., and between 25° N. and 20° S. (see Fig. 1, p. 36). This area can be specified neither by geographic boundaries nor by climatic or oceanographic regions. The chart of climatic regions by Köppen (1936) shows that it includes tropical as well as subtropical regions while according to both Schott (1935) and Dietrich (1957) it also includes several oceanic regions. Only the region of the eastern tropical Pacific Ocean lying between the California Current in the north and the Peru Current in the south is entirely covered. In the north, the southern parts of the California Current and parts of the North Equatorial Current are included, and in the south the northern portions of the Peru Current and the eastern portions of the South Equatorial Current. The California Current will not be discussed in detail in this review since extensive literature about this current is available in the Reports of the California Cooperative Oceanic Fisheries Investigations, and a general review has been given by Reid, Roden and Wyllie (1958). As regards depth only the water masses in the upper layer of the ocean, down to about 600 m depth will be discussed since, as has been shown by Reid (1965), the Intermediate Water situated near 1000 m depth requires ocean-wide considerations.

Since about 1950 interest in, and attention to, this area have increased because it is a region of comparatively high fertility, supports large fisheries, and poses interesting oceanographic problems, which have been emphasized by the discovery of the Equatorial Undercurrent and its subsequent intense investigation. Expeditions in this area using modern methods are listed chronologically in Table I. The majority of them are a result of activities of

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**TABLE I**

*List of Expeditions in the eastern equatorial Pacific Ocean*

| Name | Ship(s) | Year(s) | Source |
|---|---|---|---|
| Dana | DANA | 1928 | Carlsberg Foundation, 1937 |
| Carnegie Cruise VII | CARNEGIE | 1928–29 | Fleming, *et al.*, 1945 |
| William Scoresby | WILLIAM SCORESBY | 1931 | Discovery Committee, 1949 |
| Askoy | ASKOY | 1941 | Oceanic Observations of the Pacific |
| Albatross | ALBATROSS | 1947 | Bruneau, *et al.*, 1953 |
| California Cooperative Oceanic Fisheries Investigations | Ships of the U.S. Fish and Wildlife Service and Scripps Institution of Oceanography | 1949–64 | Oceanic Observations of the Pacific |
| Shuttle | EPCE(R)857 | 1952 | Oceanic Observations of the Pacific |
| Shellback | HORIZON | 1952 | Oceanic Observations of the Pacific |
| Hugh M. Smith Cr. 15 | HUGH M. SMITH | 1952 | Austin, 1954 |
| Acapulco Trench | SPENCER F. BAIRD | 1954 | Oceanic Observations of the Pacific |
| Eastropic | SPENCER F. BAIRD<br>HORIZON | 1955 | Oceanic Observations of the Pacific |
| | HUGH M. SMITH | 1955 | King, Austin and Doty, 1957 |
| Equapac (Hugh M. Smith Cr. 35 | HUGH M. SMITH | 1956 | Austin, 1957 |
| Scope | STRANGER | 1956 | Holmes, 1958b |
| Gulf of California | BLACK DOUGLAS<br>HORIZON<br>STRANGER | 1956–57 | Oceanic Observations of the Pacific |
| Hugh M. Smith Cr. 38 | HUGH M. SMITH | 1957 | Wilson and Rinkel, 1957 |
| Downwind | HORIZON | 1957 | Oceanic Observations of the Pacific |
| Scot | SPENCER F. BAIRD | 1958 | Holmes and Blackburn, 1960 |
| Dolphin | HORIZON | 1958 | Oceanic Observations of the Pacific |
| Doldrums | STRANGER | 1958 | Oceanic Observations of the Pacific |
| Bondy 5801–6201 | BONDY | 1958–64 | National Oceanographic Data Center |
| Tehuantepec TO 58–2 | SPENCER F. BAIRD | 1958 | Blackburn, *et al.*, 1962 |
| Tehuantepec TO 59–1 | STRANGER | 1959 | Blackburn, *et al.*, 1962 |
| Dorado | HORIZON | 1959 | Oceanic Observations of the Pacific |
| Tehuantepec TO 59–2 | HUGH M. SMITH | 1959 | Blackburn, *et al.*, 1962 |
| Costa Rica Dome | SPENCER F. BAIRD | 1959 | Oceanic Observations of the Pacific |
| Chiper | BURTON ISLAND | 1960 | National Oceanographic Data Center |
| Step-I | HORIZON | 1960 | Scripps Institution of Oceanography, Ref. 61–9, 1961 |
| Cruise 00675 | REHOBOTH | 1960 | National Oceanographic Data Center |
| Swan Song | ARGO | 1961 | On file, Scripps Institution of Oceanography |
| Shoyo Maru | SHOYO MARU | 1963–64 | Nankai Regional Fisheries Agency, Japan |

NOTE: Most of these observations have been or will be published in Oceanographic Observations of the Pacific. Many are published in data reports of the Scripps Institution of Oceanography. Observations during the International Geophysical Year are also published in IGY Oceanography Reports (Capurro, 1961).



the Scripps Institution of Oceanography. The operations of the California Cooperative Fisheries Investigation, which began in 1949 and continue to the present, extend to the southern end of the California Peninsula, and consequently the density of observations is highest in this part of the area. Since 1958 cruises of the BONDY have also resulted in an appreciable coverage of the coastal area off Peru. The Shellback Expedition in 1952 was the first to cover the whole expanse of the area under consideration, and an operation of about the same scope was repeated in 1955 during the Eastropic Expedition. The discovery of the Equatorial Undercurrent in the Central Pacific Ocean by Cromwell in 1952 led to Expeditions Dolphin in 1958 and Swan Song in 1961 to study the new phenomenon. Several smaller expeditions in the eastern tropical Pacific Ocean have investigated special problems, chiefly of a biological nature. The Step-I Expedition in 1960 covered the area of the Peru Current far beyond the coastal activities of the BONDY. In order to provide the basis for the planning of a larger systematic study of this area, a project was started in 1962 at the Scripps Institution of Oceanography and the Institute of Marine Resources to analyse all existing oceanographic observations in this area, the results of which have been incorporated into this review.

## TOPOGRAPHY

The sea floor in this area forms part of the huge Pacific Basin, and depths are in general between 3500 and 4500 m. The continental shelf is narrow and the continental slope is steep. The shelf exceeds 50 km width only along the coast of Central America between the Gulf of Tehuantepec and Nicaragua, in the Gulf of Panama, and along the coast of Peru between 7° and 11° S. The Acapulco Trench off the coast of Central America exceeds 6200 m depth near 14° N., 94° W., and, off the coast of Peru, the Peru–Chile Trench exceeds 8000 m depth at 21° S. (Fisher and Raitt, 1962).

The East Pacific Rise stretching north-south near 110° W. is crowned by a number of sea mounts and a few islands. The Galapagos Islands are connected to Central America by the Cocos Ridge extending north-east, and to Ecuador by another ridge, extending east.

The U.S. Bureau of Commercial Fisheries (1960–1962) has issued topographic charts covering this area between the coast of America and about 2000 km offshore. These charts show a large number of recently discovered sea mounts and a much more irregular topography than indicated in older maps.

## SURFACE CIRCULATION

Monthly charts of the surface currents in this area have been published by the U.S. Navy Hydrographic Office (1947) for the equatorial region between 160° W., and the coast of America and have been represented in a different fashion by Cromwell and Bennett (1959) for the area between the equator, 30° N., 120° W., and the coast of America, together with some comments about the spatial and temporal development of different branches of the circulation. For the South Pacific Ocean seasonal charts of surface currents have been published by the British Meteorological Office (1935, 1955),

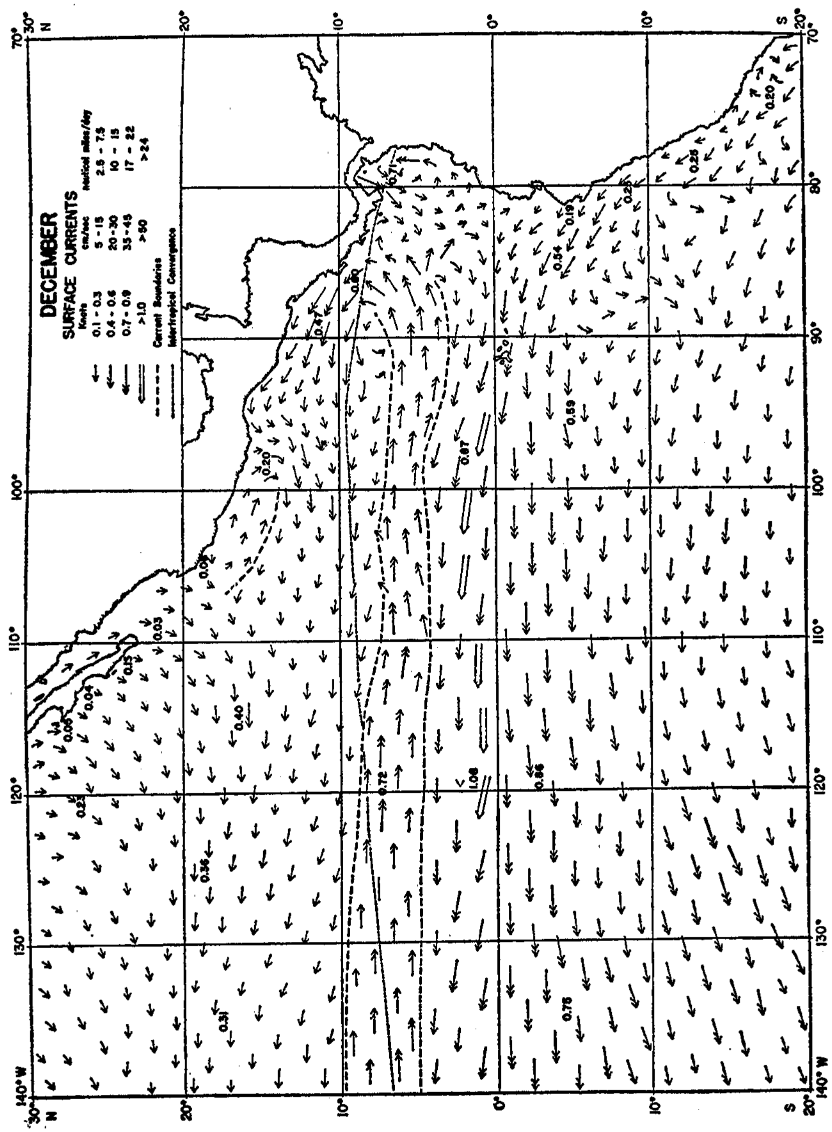


Fig. 1.—Surface currents in the eastern equatorial Pacific Ocean in December.



covering the ocean south of the equator. Both these atlases are presentations of statistically summarized observations and their interpretation is left to the reader. For many reasons, as outlined by Wyrtki (1960), interpretation of such atlases is highly desirable; the results should include the construction of current charts summarizing the data, location of the different branches of the circulation and the positions of boundaries between them, and elimination of the scattering inherent in this type of observations. With this in view, monthly

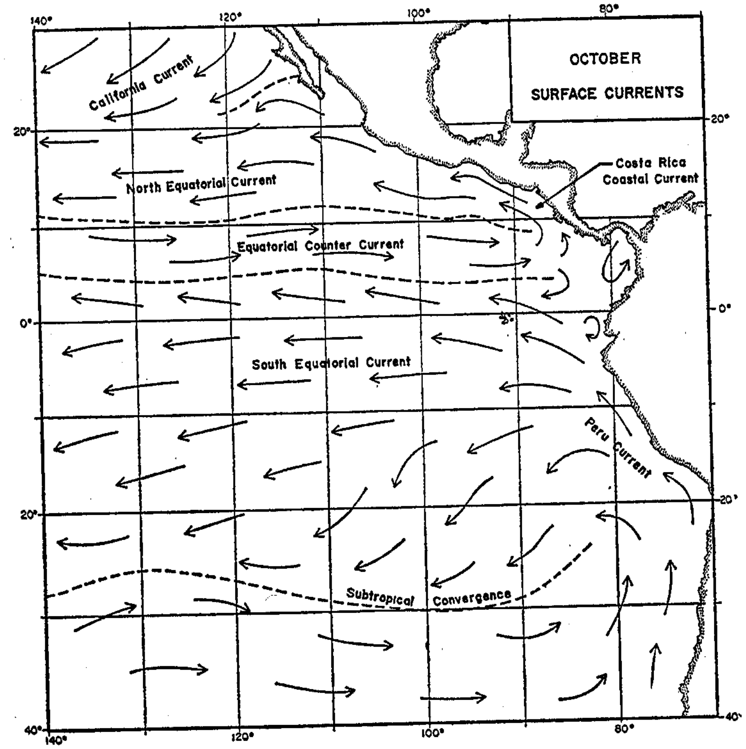


Fig. 2.—Surface circulation in October and names of the major current systems.

charts of surface currents have been constructed by Wyrtki (1965b) for the area between 30° N., 20° S., 140° W. and the coast of America. As an example the map for the month of December is reproduced in Figure 1.

According to this analysis the surface circulation in the eastern equatorial Pacific Ocean varies considerably in response to the shifting of the major wind systems (British Meteorological Office, 1956). Circulation is dominated by the eastern and equatorial parts of the anticyclonic gyrals in the North and South Pacific Ocean. In the North Pacific these consist of the California Current and the North Equatorial Current, and in the South Pacific of the Peru Current and the South Equatorial Current (Fig. 2). Between these two gyrals, the Equatorial Countercurrent is developed as long as the intertropical convergence is sufficiently far north of the equator. Because of the configuration of the ocean these two gyrals do not reach into the area of the eastern tropical Pacific Ocean between Cape Corrientes, in Mexico, and Ecuador, and consequently this area has a variable and apparently complicated circulation. This picture becomes much simpler, however, if three periods, each having a typical circulation pattern, are compared with each other.

The first typical circulation pattern is developed from August to December when the Equatorial Countercurrent is fully developed and the South Equatorial Current is also very strong, especially in that part situated north of the equator. During this period most of the water of the Countercurrent flows around the Costa Rica Dome, which is situated near 9° N., 89° W., into the Costa Rica Coastal Current and enters the North Equatorial Current between 10° N. and 20° N. The California Current leaves the coast off Lower California at about 25° N. and supplies the water of the north Equatorial Current only north of 20° N. This situation is developed when the inter-tropical convergence is in a northerly position at, or north of 10° N., approximately coinciding with the northern boundary of the Countercurrent. This circulation pattern is the most stable and lasts longest. In January, when the inter-tropical convergence starts to move towards the equator, the Countercurrent becomes much weaker and breaks up into several segments. Simultaneously the California Current becomes stronger and extends farther to the south.

The second typical circulation pattern is developed from February to April, when the inter-tropical convergence is in its most southerly position near 3° N. During this period the California Current is strong, penetrates far to the south, and supplies most of the water for the North Equatorial Current. Off the coast of Mexico between 10° N. and 20° N., the circulation is anticyclonic with a flow to the southeast along the coast which turns westwards off the Gulf of Tehuantepec. The Countercurrent is absent during this period, as shown by Knauss (1958), and water flows west and northwest where the countercurrent is otherwise found. Off the coast of Central America two huge eddies are developed, one is cyclonic around the Costa Rica Dome, the other anticyclonic around a point at 5° N., 88° W. The South Equatorial Current is weaker, and currents to the east are occasionally reported near the equator. The Peru Current is also relatively weak, and the Peru Countercurrent is pronounced.

The third typical circulation pattern is developed from May to July, when the Countercurrent forms again, but the California Current is still relatively strong. During this period the inter-tropical convergence is again near 10° N., which allows the Countercurrent to develop. Most of the water from the Countercurrent turns north into the Costa Rica Coastal Current, which, during this period, flows along the coast of Central America as far as Cape Corrientes. The California Current is still strong and reaches far to the south, but does not penetrate much into the eastern tropical Pacific. It forms the main supply of the North Equatorial Current. From July to August the California Current becomes progressively weaker, and the North Equatorial Current gains more and more water from the eastern tropical Pacific.

This surface circulation is reflected in the topography of the mixed layer as first pointed out by Cromwell (1958). The close relationship between circulation and the topography of the mixed layer and the depth of the centre of the thermocline is shown in monthly charts prepared by Wyrtki



(1965a). The map of the geostrophic flow at the surface of the Pacific Ocean by Reid (1961) also shows all these main features of the surface circulation. Based on geostrophic calculations, Reid (1959, 1961) and Wooster (1961a) postulate the existence of a South Equatorial Countercurrent between 5° S. and 8° S. This current is likely to be more pronounced in sub-surface layers, and it may be concealed by the wind drift at the surface, since, as stated by Wyrtki (1965b), no surface current observations to the east near its tentative position have been reported.

At the equator, west of the Galapagos Islands, occasional surface flow to the east has been reported and has been charted by Puls (1895) and Schott (1935). These observations are regarded by Knauss (1960) as a surfacing of the Undercurrent in the absence of easterly winds, and have been explained theoretically by Roden (1962).

Off the coast of Peru the flow is to the northwest as part of the anticyclonic circulation in the South Pacific, and the water of the Peru Current passes into the South Equatorial Current. Along the coast, coinciding approximately with the upwelling area, the Peru Coastal Current flows northwest, and farther offshore the Peru Oceanic Current also flows northwest (Gunther, 1936). These two currents are usually separated by a weak and irregular flow to the south, namely, the Peru Countercurrent (Wyrtki, 1963). This current is, however, a sub-surface current and only occasionally reaches the surface. From July to October the Countercurrent is not observed at the surface; from November to March it is most pronounced and is situated about 500 km offshore; and from April to June southward flow is only occasionally observed and its location is more variable.

## SURFACE TEMPERATURE AND SALINITY

### TEMPERATURE

In recent years there have been many additions to the data concerning the distribution of sea surface temperature, as summarized in atlases of the British Meteorological Office (1956) and of the U.S. Navy Hydrographic Office (1944, 1960). Monthly charts of sea surface temperature for this area have been issued since 1960 by the Bureau of Commercial Fisheries, San Diego, (Johnson, Flittner and Cline, 1965) and average monthly charts for the coastal regions as well as monthly anomaly charts as far back as 1947 have been published by Renner (1963). Using all available data, Wyrtki (1965a) has published monthly charts of sea surface temperature (Fig. 3) and charts showing the change of surface temperature from month to month. These charts are in general similar to those of the two atlases, but certain small-scale features now to be described, appear more clearly. The tongue of warm water situated off the coast of Peru from October to May is confirmed (Fig. 3); this tongue is not shown at all by Schott (1935), and the atlas of the U.S. Navy Hydrographic Office (1944) gives much lower temperatures. The warm water situated off Central America is not as homogeneous as previously assumed and the effects of upwelling are clearly seen in the Gulf of Tehuantepec from November through March, and in the Gulf of Panama from February through April. Further, the upwelling in the Costa Rica Dome is indicated by lower surface temperatures from December through May.

Results of an harmonic analysis of sea surface temperatures for the Pacific Ocean north of 20° S. published by Wyrtki (1965c) show that in the eastern equatorial Pacific Ocean a continuous change of the phase of the annual temperature wave takes place from the southern to the northern hemisphere. Saur (1963) has critically discussed the accuracy of surface temperature taken by thermometers in the sea water intake of ships. A more detailed presentation of sea surface temperatures for the Gulf of Panama has been given by Sund (1963). For the years 1956 and 1957 monthly charts of sea surface

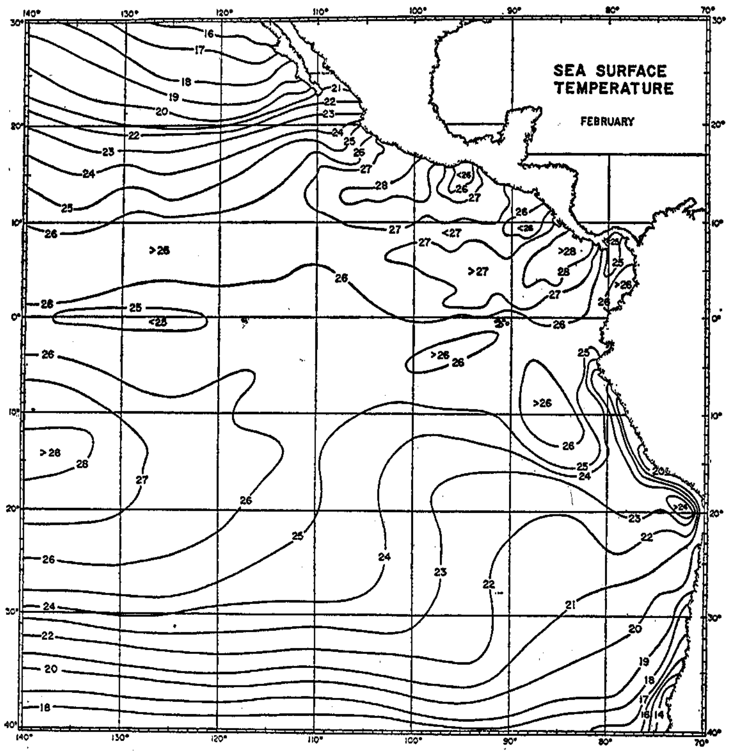


Fig. 3.—Sea surface temperature in the eastern Pacific Ocean in February.

temperature for the Pacific Ocean north of 20° S. have been published by the U.S. Bureau of Commercial Fisheries, Stanford (1962).

## SALINITY

The charts of surface salinity drawn by Schott (1935) have for a long time been the only available representation. Recently, Bennett (1965) has constructed monthly maps of surface salinity for this area, indicating also the areas without coverage. Although these maps are not yet a final representation of surface salinity and many more observations are required for this, they do show some characteristic features of the seasonal changes. The seasonal



variation seems to be relatively weak in the South Pacific and in the range of the California Current. In the eastern tropical Pacific Ocean seasonal changes are much larger, especially in the vicinity of the Gulf of Panama. Lowest salinities of less than 30‰ are found from October through January. Later on this water of low salinities moves west, whereby salinity in its centre increases. From the coast of Central America an area with salinities of less than 34‰ extends west near 10° N. It shifts slightly in latitude with the season, being displaced south from April through June and north during the second half of the year. The high salinity water formed in the Gulf of California is connected during the entire year with the high salinity water of the sub-tropical North Pacific Ocean by a ridge with salinities between 34·0‰ and 34·5‰. This ridge separates the water of low salinity of the California Current from the Tropical Surface Water also having low salinities.

## THERMAL STRUCTURE

In these tropical and sub-tropical parts of the ocean the thermal structure is characterized by a mixed layer, in which temperature is almost constant, by

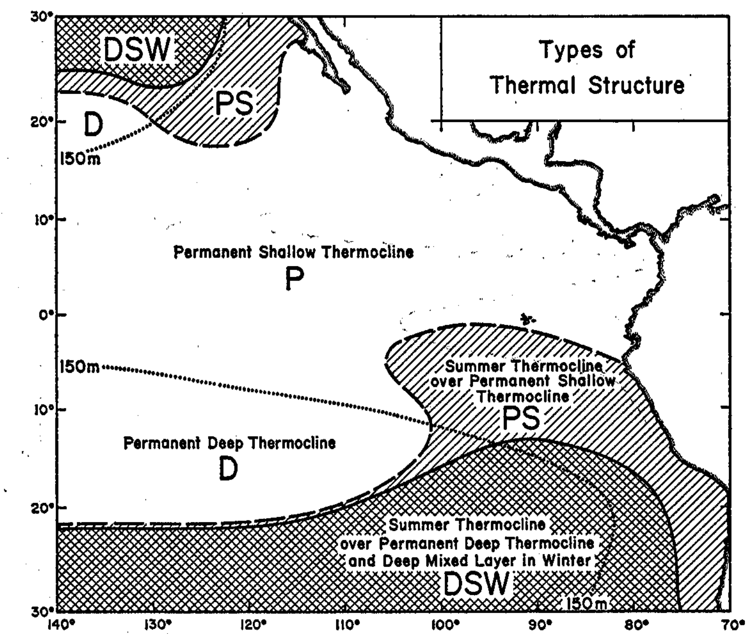


Fig. 4.—Distribution of types of thermal structure in the eastern equatorial Pacific Ocean.

the thermocline in which temperature decreases sharply and substantially, and by a sub-surface layer in which temperature continues to decrease, but much more slowly. Cromwell (1958) was the first to chart the topography of the mixed layer for the four seasons, comparing it with circulation and drawing conclusions about enrichment processes connected with ascending movements. After substantially more data were accumulated, Wyrtki (1965a) published monthly charts of the depth of the mixed layer, and of the depth of the centre of the thermocline. In the sub-tropical parts of the region where the seasonal variation of temperature is more pronounced, a summer thermocline forms on top of the permanent thermocline.

Based on the analysis of these charts and on diagrams of the seasonal changes of the thermal structure, characteristic types can be distinguished

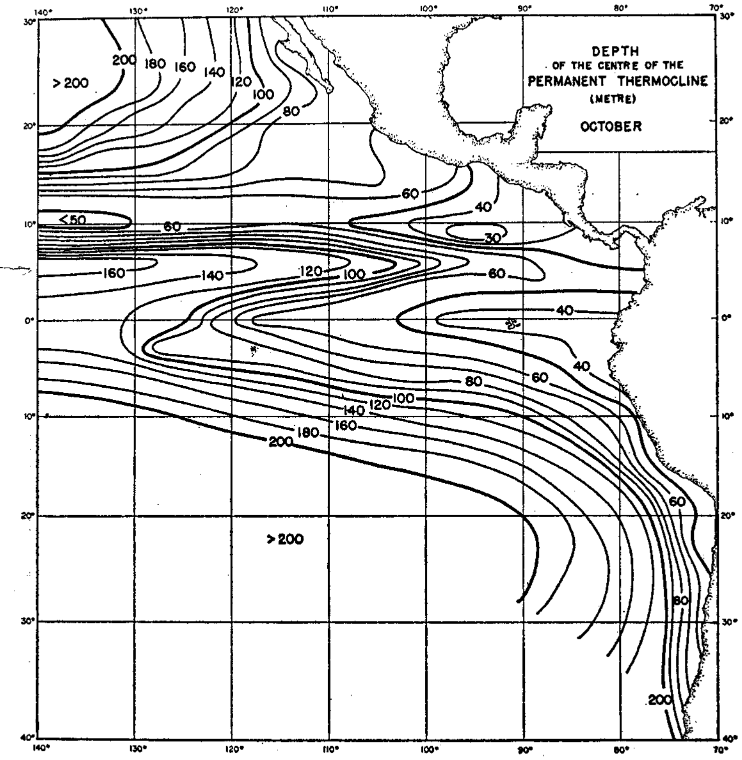


Fig. 5.—Depth of the centre of the permanent thermocline in the eastern equatorial Pacific Ocean in October.

and their distributions charted (Fig. 4). A permanent shallow thermocline is characteristic for the eastern tropical Pacific, which is bordered by the California Current in the north and by the Peru Current in the south, both having a different thermal structure. In this region the temperature gradient in the thermocline usually exceeds 1° per 10 m and the difference of temperature between the top and the bottom of the thermocline is large. West of 130° in the northern hemisphere, and west of 100° in the southern hemisphere, regions with a permanent deep thermocline appear on both sides of the region having a permanent shallow thermocline. The former are typical for the northern portions of the North Equatorial Current and the southern portions of the South Equatorial Current. In general, a permanent shallow thermocline



characterizes the inner tropical region, and a permanent deep thermocline the outer tropical region. The sub-tropical region, where the seasonal variations in surface temperature become appreciable, is characterized by the development of an additional summer thermocline. The temperature gradient is usually stronger in this summer thermocline than in the deeper permanent thermocline. Where such a summer thermocline is formed, the mixed layer is deep in winter and spring, and shallow in summer and autumn. This type of thermal structure is chiefly found in the southern portions of the California Current and in the northern portions of the Peru Current, from where it extends south of the Galapagos Islands to about 105° W. The geographical distribution of the different types of thermal structure is shown in Figure 4 and the topography of the centre of the permanent thermocline in Figure 5.

In the equatorial region the topography of the mixed layer is characterized by a sequence of ridges and troughs. Along the equator the mixed layer is very shallow during the entire year due to equatorial upwelling related to the Undercurrent (Austin, 1960). A trough near 4° N. and a ridge near 10° N. indicate the southern and northern boundaries of the Countercurrent. While the position of the trough is very stationary, that of the ridge exhibits noticeable north-south fluctuations, which are related to the strength of the Countercurrent. Also the depths of the mixed layer and of the thermocline along the ridge and the trough vary considerably during the year and are related to a tilting of the discontinuity layer beneath the Countercurrent. From January to June another trough is situated further north, near 13° N. In indicates the position where the water flowing southeast off the coast of Mexico turns west into the North Equatorial Current. Within the North and South Equatorial Currents the depth of the permanent thermocline increases with increasing distance from the equator, whereas the depth of the mixed layer is dependent on the development of the summer thermocline.

The relations of the topography of the thermocline to the transverse circulation in ocean currents and to ascending and descending movements have been widely discussed and will be treated separately (see p. 57).

## OCEANIC FRONTS

One of the most pronounced fronts in lower latitudes is the equatorial front stretching from the coast of Ecuador at about 4° S. towards the Galapagos Islands (see Fig. 7, p. 48). This front separates the tropical water of high temperature and low salinity in the north from the cooler water of high salinity of the Peru Current. It indicates the line in which the upper parts of the extremely strong tropical thermocline reach the sea surface. From about May to November this front is characterized by a very sharp contrast of temperature and salinity, the temperature difference being sometimes 5°–6° C and the salinity difference 1‰. From January to March when the water south of the equator has temperatures similar to those in the north, the front is still marked by a substantial salinity difference. This front extends from the coast of Ecuador to the north-west, cuts the equator to the east of the Galapagos Islands, and continues between 1° N. and 3° N. to the west. It is strongest near the coast and becomes progressively weaker towards the west, where mixing destroys its structure.

Bjerknes (1961) has pointed out that the warmer water north of the front has a tendency to flow over the cooler water to the south according to the pressure gradient at the surface. In sub-surface layers the cooler water flows north. As a result of these movements the front would advance south, as has been observed, unless southerly winds keep it in its position. With this convective circulation in mind, Bjerknes (1961) argues that a disappearance of southerly winds for an appreciable time would result in an overflow of warm tropical water over the cooler water of the Peru Current producing a situation commonly known as El Niño; Fedorov (1963) also discusses trans-equatorial circulation in the eastern Pacific and comes to the same qualitative conclusions. For estimating the velocity of the meridional movements, he used a method based on friction which is superior to the estimates by Bjerknes (1961) based on the acceleration during one day.

West of about 100° W. the equatorial front is situated between about 2° N. and 4° N. and runs parallel to the equator (see Fig. 7, p. 48). There it is chiefly characterized by a discontinuity in temperature and much less in salinity. This front has been observed many times and is described in detail by Cromwell and Reid (1956) for crossings at 120° W. and 172° W. and by Knauss (1957) for another crossing at 120° W. Cromwell and Reid (1956) discuss the transverse circulation at such a front and conclude that surface flow is convergent; sinking of water occurs at the front and sub-surface flow at the top of the main thermocline is divergent. This circulation scheme would maintain the front and dispose of the mixed water formed along the front by the sinking and the divergent flow below the front. Moreover, they suggest that vertical mixing should be different on both sides of the front. Knauss (1957) discusses a completely different situation along a front. The frontal zone is moving under the influence of winds from the colder water to the warmer water, and colder water overflows the front at the surface, creating an unstable zone of strong mixing, about 100 m wide. It seems that the first case is characteristic for the front along the southern boundary of the Countercurrent, where convergent movements can be expected. The second case might be more representative for a situation in which the warm water is prevented from flowing over the cooler water by an opposing wind and, as discussed above, is probably characteristic of the front between Ecuador and the Galapagos Islands.

It is not yet established whether the equatorial front to the west of about 100° W. is a continuation of the front stretching from Ecuador to the Galapagos Islands. It may not develop during all seasons, is certainly subject to some lateral movements, and may vary considerably in intensity. It is also very likely that the front is not continuous, but consists of individual segments. It is not expected that a similar front will develop along the northern boundary of the Countercurrent because movements there are divergent and unfavourable to its formation.

Between the water of the California Current and that of the Gulf of California another front, starting at Cape San Lucas at the southern tip of the California Peninsula, develops (Griffiths, 1965). The difference in density at this front is small because cool low salinity water of the California Current meets warm water of higher salinity from the Gulf. This front is well developed near the Cape but seems to fade out rapidly towards the south-west, probably because the Gulf of California can supply only small amounts of water. Detailed measurements of the temperature structure off the California Peninsula and south of Cape San Lucas have recently been made using a thermistor chain (LaFond, 1963). A power spectrum analysis of these records shows that four waves with wavelengths between 0·5 and 1·6 km are more emphasized than others. Examples of various thermal features as fronts, inversions and ridges are shown.



## HEAT EXCHANGE AT THE SEA SURFACE

The components of the heat exchange at the sea surface have been calculated from climatic data by Budyko (1955, 1956) for the entire earth, but his charts show little detail for the region under consideration. A re-calculation of the heat exchange (Wyrtki, 1965c, 1966), using two-degree-squares and months as

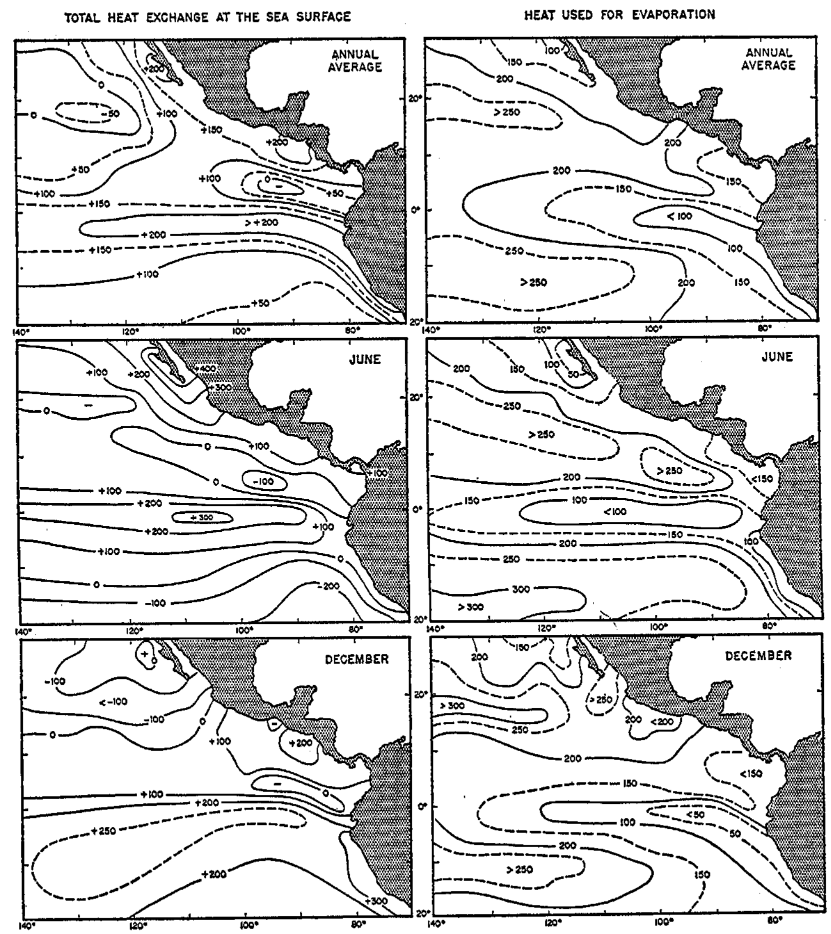


Fig. 6.—Distribution of the total heat exchange at the sea surface and of the energy used for evaporation in the eastern equatorial Pacific Ocean for June, December, and annual average in cal cm$^{-2}$ day$^{-1}$.

units, has been made for the North Pacific Ocean including this area, and the more detailed distribution obtained is closely related to climatic regions. In particular, zonal features of small meridional extent not even shown in Budyko's atlas are much more pronounced in the Wyrtki's analysis. Maps of the distribution of the average annual heat exchange at the sea surface and of the average annual heat loss due to evaporation, as well as the distribution of these quantities in June and December, are shown in Figure 6.

Most of the area gains heat during the year, as would be expected in tropical and sub-tropical regions. The area of heat loss between 15° N. and 20° N. off Lower California is caused by high evaporation under the north-east trades. In winter this heat loss exceeds 100 cal $cm^{-2}day^{-1}$. Just south of the equator there is a belt of considerable heat gain, due chiefly to reduced evaporation in this area. This belt also extends along the coast of Peru, where the gain of heat is compensated by the advection of cooler water in the Peru Current and by the upwelling along the coast. The average heat gain in the eastern tropical Pacific Ocean is probably balanced by horizontal diffusion of heat out of the region and by ascending of cooler sub-surface water. The interaction between the heat exchange at the sea surface, the thermal structure and circulation have not yet been studied for this region.

## WATER MASSES

Before 1950 only a few hydrographic stations had been occupied in this area; the results of all the early expeditions have been summarized by Wooster and Cromwell (1958). In 1952 the Shellback Expedition covered most of the area under consideration for the first time. The results of this expedition, published by Wooster and Cromwell (1958), show the distributions of temperature, oxygen content, phosphate, and silicate along all the sections of this expedition. Salinity samples of the expedition were unusable because bottles were not adequately sealed and evaporation took place. Several features of interest in these sections are described in the report, but no analysis of water masses is given probably because salinity data were lacking. The authors comment on upwelling west of the Galapagos Islands and in the Costa Rica Dome, on enrichment processes in the Peru Current and in the eastern portions of the South Equatorial Current, and on the front between Ecuador and the Galapagos Islands. But more important, the authors document comprehensively the extent of the oxygen minimum layer in this area, and demonstrate that waters of high nutrient content are present below the shallow mixed layer.

The next comprehensive coverage of the area was achieved in 1955 by the Eastropic Expedition and the results have been analyzed by Bennett (1963). The distribution of properties is shown not only along sections but also on three surfaces of constant thermosteric anomaly; however, no interpretation of this material is included.

The first hydrographic section along the equator from 140° W. to the coast of America was obtained during the Dolphin Expedition in 1958. Other expeditions have either restricted their activities to smaller areas, such as the Gulf of Tehuantepec, the Gulf of California, or the Costa Rica Dome, or have produced data scattered over wide areas, as did the Downwind Expedition.



The Step-I Expedition in 1960 investigated the area off Peru, discovering and documenting an extensive subsurface flow to the south (Wooster and Gilmartin, 1961). They show that southward spreading of water of low oxygen content and high salinity near 200 to 300 m depth is related to this sub-surface flow. An analysis of the water masses in this region has been given by Wyrtki (1963) in connection with a discussion of the field of motion in the Peru Current.

The only existing discussion of the distribution of the main water masses in this area is still that given for the entire Pacific by Sverdrup, Johnson and Fleming (1946), but their analysis chiefly concerns sub-surface water with temperatures between 8° C and 16° C and almost the entire area under consideration would be classified as equatorial Pacific water. It seems appropriate, therefore, in this review to give an analysis of the water masses in this region although the results are not yet published (Wyrtki, in press).

### SURFACE WATER MASSES

At the surface there are three basically different types of water involved: (1) tropical surface water of high temperature and low salinity, (2) sub-tropical surface water of high salinity, which is generally warm but variable in temperature, and (3) surface water of the California and Peru Currents, which is cool and of low salinity and originates in higher latitudes (Fig. 7). Naturally, all boundaries between these water masses are subject to seasonal fluctuations, and in most cases they are boundary zones rather than fronts.

*Tropical Surface Water* is found in regions where sea surface temperature is high and its seasonal variation small, and where salinity is low due to an excess of rainfall over evaporation. In the eastern tropical Pacific this water can be identified by the area where surface temperature is always above 25° C. Within this area, salinity is usually less than 34‰ due to an excess of rainfall over evaporation which, according to Dietrich (1957), is greater than 50 cm/year. The southern boundary of tropical surface water runs from Ecuador to north of the Galapagos Islands and continues west at about 4° N. where it coincides approximately with the southern boundary of the Countercurrent. The water carried east with the Countercurrent as well as that carried west in the southern parts of the North Equatorial Current is tropical surface water. The northern boundary of the tropical surface water can be identified approximately with the 25° C isotherm which lies near 15° N. and fluctuates during the year by about 5° of latitude. Lowest salinities within this water are found in the Gulf of Panama and off the coast of Columbia, where salinity varies from 34‰ to less than 30‰ at the end of the rainy season (Bennett, 1965). Except along the southern boundary of the Countercurrent, where it can be as much as 100 m deep, the vertical extent of the water is limited to the shallow mixed layer, usually only 20–50 m thick. Temperature decreases and salinity increases within the sharp discontinuity layer below this mixed layer.

The *Sub-tropical Surface Water* of the South Pacific Ocean is formed in the regions where evaporation greatly exceeds precipitation. It is characterized by high salinity, but temperature in this water mass can vary over a wide range from about 28° to 15° C. In the South Pacific Ocean the highest surface salinities are found between 12° S. and 25° S., and between 100° W. and

150° W. where salinity is above 36‰. The centre of the sub-tropical surface water coincides approximately with the centre of the South Pacific anti-cyclone. The long residence time of the surface water near its centre, where evaporation exceeds precipitation by about 100 cm/year (Dietrich, 1957), allows the salinity to be raised to these high values. In the centre of the sub-

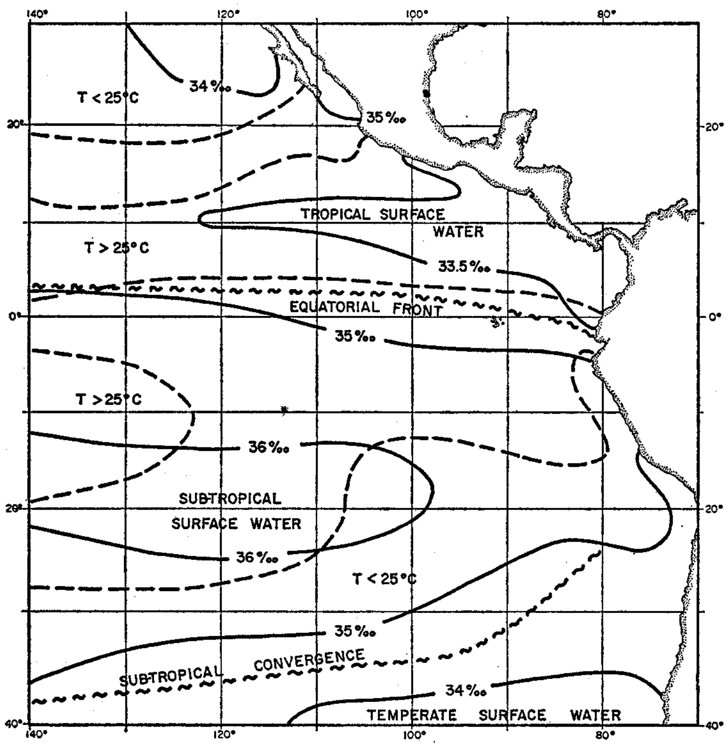


Fig. 7.—Distribution of the main surface water masses in the eastern Pacific Ocean: ——— surface salinity; - - - - extreme positions of the 25° C isotherm: ~~~~ major oceanic fronts.

tropical anticyclone near 20° S., salinities above 36‰ are found as deep as 200 m—a result of the deep-reaching convection in winter when the mixed layer extends almost to that depth. In summer, a shallow thermocline is found near the surface. The eastern and northern boundary of the sub-tropical surface water cannot be precisely determined; off Peru this water is often separated from the coast by only a narrow belt of upwelling water of lower salinity and temperature; the northern boundary of this water can be identified approximately with the 35‰ isohaline which, starting from the coast of Ecuador at about 5° S., runs south of the Galapagos Islands to the west and continues along the equator.

This northern boundary of the sub-tropical water of the South Pacific Ocean does not coincide with the southern boundary of the tropical surface water. Between these two water masses there is another water mass, which,



being intermediate in properties, will be called *Equatorial Surface Water*. This equatorial surface water is not a plain mixing product of the two other water masses, but its properties are determined by seasonal advection of cooler water from the Peru Current and by equatorial upwelling. This water is typical for those parts of the South Equatorial Current situated at and north of the equator.

The *Water of the California Current* is of moderate temperature and low salinity and flows south. Where this water turns away from the coast off lower California to form part of the North Equatorial Current, temperature and salinity continue to increase until the water of the California Current is converted into sub-tropical surface water of the North Pacific Ocean. Off the southern part of the California Peninsula the water of the California Current meets the tropical surface water, but no permanent sharp front is developed. Since surface salinity increases towards the south in the California Current and decreases again in the eastern tropical Pacific, the boundary between these two water masses is indicated by a weak salinity maximum at the sea surface, and this connects the high salinity water of the Gulf of California with the sub-tropical surface water farther west (Bennett, 1965). Where the water coming from the California Current and the tropical surface water coming from the eastern tropical Pacific Ocean jointly form the North Equatorial Current no obvious boundary exists between the tropical and sub-tropical waters, but a gradual transition of properties takes place.

In the *Gulf of California*, water of high salinity is formed due to an excess of evaporation over rainfall; this water can be classified as sub-tropical. Salinity in the Gulf is above 35‰ and occasionally in some places it can reach 36‰; temperature in the Gulf varies considerably from 15° C to 30° C (Roden and Groves, 1959). This high salinity water leaves the Gulf at the south and spreads, according to its temperature, either at the sea surface or as a sub-surface salinity maximum. The amount of this water of high salinity formed in the Gulf is so small that it exerts little influence outside the Gulf itself.

## SUB-SURFACE WATER MASSES

Sub-surface water masses formed in the region are the Sub-tropical Sub-surface Water of high salinity and the layer of the oxygen minimum. All deeper water masses are formed outside the region and penetrate into the area under consideration by horizontal flow and large scale horizontal mixing. This is the case for the Intermediate Water characterized by a salinity minimum at 700 to 900 m depth and which has been comprehensively discussed by Reid (1965), and also for the Pacific Deep Water as discussed by Wooster and Volkman (1960) and by Knauss (1962). Neither of these water masses will be discussed here, because their treatment would require ocean-wide consideration.

The *Sub-tropical Sub-surface Water* originates in the sub-tropical South Pacific Ocean, where surface water of high salinity is formed. During winter, the surface layer of high salinity and relatively low temperature is homogeneous down to depths of more than 150 m in the central parts of the sub-tropical anticyclone. Montgomery (1959) has calculated the residence time of the Sub-tropical Surface Water in the areas of its formation and finds that in the South Pacific Ocean the residence time is approximately 12 years. From there a sub-surface salinity maximum extends towards the equator underneath the surface water of lower salinity. This salinity maximum is situated near to 100 m depth in the upper portions of the thermocline. The flow in these depths is to the west with the South Equatorial Current, but a slight component towards the equator is superimposed. When approaching the equator, the salinity maximum layer comes under the influence of the Equatorial Undercurrent and the water of high salinity is carried east. Only the lower portions of the salinity maximum layer and the mixed water discharged by the

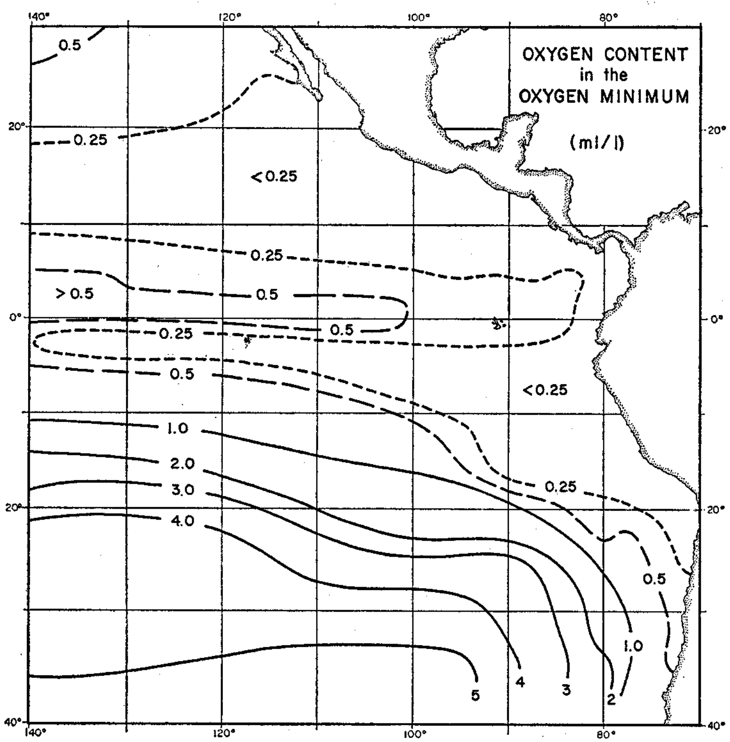


Fig. 8.—Oxygen content in the oxygen minimum situated near 400–500 m depth.

Equatorial Undercurrent reach the northern hemisphere; there, the salinity maximum is at a deeper level with a lower temperature and with a much reduced oxygen content. This salinity maximum is found in the entire eastern tropical Pacific underneath the surface water of low salinity and extends to about 20° N. Slow ascending movements bring this water from depths between 100 and 150 m into the surface layer where it mixes with the surface water, the salinity of which is continuously reduced by the excess of rainfall over precipitation in this region; the velocity of these ascending movements has been calculated as 20 m/year. The salinity maximum layer suffers a loss of water at the rate of about 5 million $m^3$/sec and the average residence time of the water in this layer is about 10 years.

An *oxygen minimum layer* is present throughout the entire region, even



though its oxygen content varies considerably (Fig. 8). The layer in which the oxygen content is less than 1 ml/l is more than 1200 m thick off the coast of Mexico and more than 800 m thick off Peru (Fig. 9). Along the equator its thickness is less than 300 m, thus dividing the two huge bodies of water with extremely low oxygen content. Within the central parts of these water bodies, oxygen content becomes less than 0·25 ml/l but no formation of $H_2S$ has

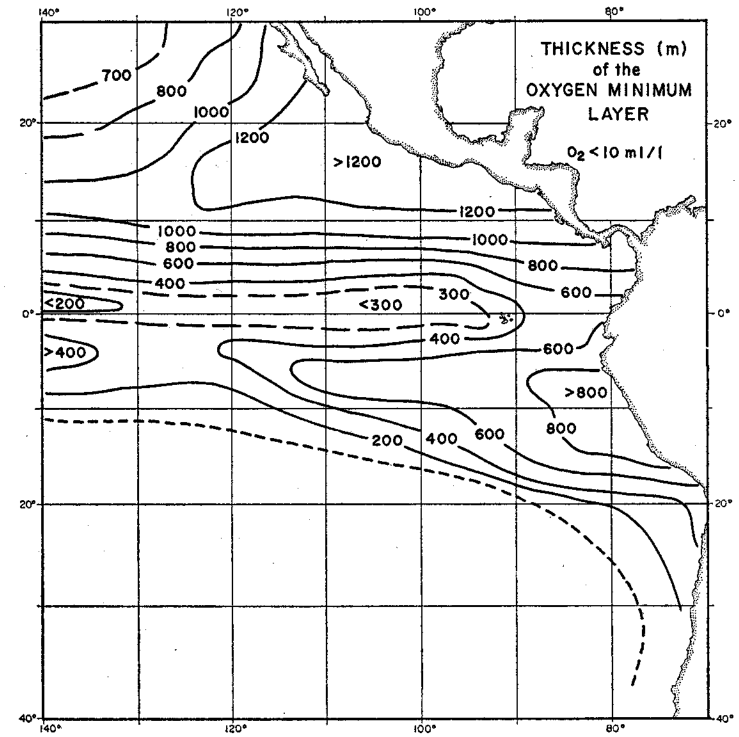


Fig. 9.—Thickness of the oxygen minimum layer, where the oxygen content is less than 1·0 ml/l.

been observed. The actual minimum lies between 300 and 500 m depth. The upper boundary of this oxygen minimum layer comes to within 50 m of the sea surface off Central America and off Peru. Along the equator the upper boundary is deeper, at 250 m, due to the influence of the Equatorial Undercurrent.

In each hemisphere such oxygen minima seem to develop at these levels in the eastern equatorial corners of the ocean basins into which the anticyclonic circulation does not penetrate. Because of the sluggish circulation the residence time of the water in these strata is long, and the consumption of oxygen is high due to the high productivity in the surface layer over large parts of the region (Reid, 1962). In the absence of a strong circulation, oxygen is supplied to the oxygen minimum layer only by vertical and horizontal diffusion, and by water ascending from deeper levels (Wyrtki, 1962). The oxygen minimum off Mexico is more extensive than that off Peru because in general the North Pacific has a lower oxygen content.

Most of the water discharged by the Equatorial Undercurrent near the Galapagos Islands at 100 to 200 m seems to penetrate into the oxygen minimum layers to the north and to the south, and so supplies these layers with oxygen and also with water of relatively high salinity (above 34·9‰). Moreover a salinity maximum is situated above the oxygen minimum layer from which, by vertical diffusion, the salinity in the oxygen minimum layer is increased. On both sides of the equator the oxygen minimum layer is therefore of relatively high salinity.

From the oxygen minimum layer off Peru an oxygen minimum coinciding with a salinity maximum extends to the south along the coast of Chile indicating that there is a flow of water to the south in sub-surface layers between 200 and 400 m depth. The spreading of this water to the south is accomplished by the Peru Countercurrent (Wooster and Gilmartin, 1961; Wyrtki, 1963).

An *Upper Salinity Minimum* occurs off the coast of Chile and off the coast of California. South of the Sub-tropical Convergence there is water with a salinity of less than 34‰ and a temperature below 15° C and this flows north off the coast of Chile in the eastern portions of the anticyclonic gyral in the South Pacific. At about 35° S., in the vicinity of the Sub-tropical Convergence, this water slides underneath the Sub-tropical Water which has slightly higher salinities and temperatures, but a lower density. The spreading of this Sub-antarctic Water can be followed by a salinity minimum at 100 m depth stretching north along the coast of Chile and Peru to about 15° S. It extends west to about 95° W. where the salinity minimum is found at more than 250 m. Along the coast of southern Peru, south of 15° S., this water comes very close to the surface and supplies the upwelling current.

## CIRCULATION

### EQUATORIAL UNDERCURRENT

The discovery of the Equatorial Undercurrent in 1952 by Cromwell completely changed the previously existing concept of the equatorial circulation, and it seems justified, therefore, to start this section with considerations of this current, in spite of the fact that it is entirely below the sea surface. After the first observations concerning this current had been reported by Cromwell, Montgomery and Stroup (1954), measurements were made during the Eastropic Expedition in 1955, (King, Austin and Doty, 1957) and in 1958 the Dolphin Expedition (Knauss and King, 1958) was designed to investigate the current between 140° W. and the Galapagos Islands. Based chiefly on observations during this expedition Knauss (1960) has described the physical characteristics of the current. The Undercurrent is a sub-surface current with maximal velocities of 120–150 cm/sec at the equator at about 100 m depth. The current is symmetrical with respect to the equator, it is about 200 m thick and 300 km wide and its transport lies between 34 and 42 . $10^{12}$ $cm^3$/sec. Above the Undercurrent a thin layer of about 20 to 50 m thickness moves west. The current remained virtually unchanged over a period of two weeks and extended all the way from 140° W. to the Galapagos Islands, but decreased in speed close to these Islands; east of the Islands it was not found. Isotherms in a section across the equator show spreading, and weaker vertical temperature gradients at the equator coinciding with the position of the Undercurrent and the isolines of other properties are distorted in a corresponding fashion. Austin (1958) describes the distribution of properties along the equator and finds that the zonal pressure gradient extends only to approximately 300 m depth.



Knauss (1960) states that zonal motion in the Undercurrent is in approximate geostrophic balance (this has been confirmed by Montgomery and Stroup, 1962), and he explains the meridional pressure gradients as a result of intense mixing near the core of the Undercurrent. However, as he states, this does not explain the existence of the current, and he points out that the Undercurrent might be produced by the wind-induced divergence at the equator. This divergence and the resulting equatorial upwelling had been discussed by Cromwell (1953) before the discovery of the Undercurrent.

The discovery of the Undercurrent was followed by a number of articles attempting to explain it in theoretical terms, although no theory had, in the past, indicated or postulated its existence! Fofonoff and Montgomery (1955) examined the vorticity equation and found that the Undercurrent is consistent with the conservation of absolute vorticity and that a zonal pressure gradient is essential for the existence of the current. They construct a scheme of meridional circulation, which requires a divergence at the sea surface at the equator, upwelling at the equator and a convergence in the discontinuity layer, coinciding with the core of the Undercurrent. The meridional velocity towards the equator is calculated as 4 cm/sec. Stommel (1960) has examined the equation of motion and finds that non-linear terms cannot be neglected. Charney (1960) points out that inertial forces are equally important as frictional forces within the Undercurrent, and he also proposes a scheme of meridional circulation, which in the upper portion is identical with that proposed by Cromwell (1953) and by Fofonoff and Montgomery (1955), but includes downward movements below the core of the Undercurrent. Veronis (1960) has calculated the vertical distribution of zonal velocity from the zonal pressure gradient. Wyrtki and Bennett (1963) use Bernoulli's equation to calculate velocities in the core of the Undercurrent, and find that the observed distribution can be explained by a vertical eddy viscosity of A = 3 $cm^2$/sec. Close to the Galapagos Islands the velocity decreases, due to the combined action of friction and a retarding pressure gradient. Montgomery and Stroup (1962) in discussing a very detailed section across the equator taken in 1952, found that the Undercurrent is in geostrophic balance, and they proceed to determine the amount of water of certain temperature-salinity combinations transported by the Current. The main flux of the Undercurrent at 150° W. consists of water with temperatures near 13° C and salinities near 34·9‰. Stroup (1961) discusses the distribution of this 13° C water in the eastern Pacific Ocean, shows that its thickness increases to the east and that near the Galapagos Islands it spreads north into the eastern tropical Pacific and south under the Peru Current. Montgomery and Stroup (1962) calculated the total transport of the Undercurrent as 34 . $10^{12}$ $cm^3$/sec. Direct observations of currents at the equator have been reviewed by Montgomery (1962), and a general account of equatorial current systems has been given by Knauss (1963). In 1961 Expedition Swan Song again observed the Undercurrent on several north-south sections, and gave special attention to meridional movements, but the results have not yet been published in detail.

Since the Equatorial Undercurrent has considerable effects on the distribution of water masses in the eastern equatorial Pacific, its discovery has changed the concept of the circulation, and many previously unexplained phenomena can now be understood. According to the transverse circulation described by Charney (1960), the Undercurrent draws water out of the range of the discontinuity layer towards its core. This water is subject to considerable mixing in the core, as is evident from the near disappearance of the salinity maximum of the Sub-tropical Sub-surface Water at the equator. Above the core of the Undercurrent this mixing water ascends into the surface layer and causes the lower surface temperatures at the equator, associated with upwelling. This upwelling is intensified by the divergence of the wind-induced flow at the surface. Below the core of the Undercurrent, mixing water is depressed; both effects result in the characteristic spreading of the isotherms near the equator. Austin (1960) discusses the strong advection of water of high oxygen content from west to east by the Undercurrent.

Near the Galapagos Islands, where the strength of the Undercurrent decreases, water probably turns north and south, as suggested by Knauss (1960) and it is very likely that parts of this water supply the sub-surface salinity maximum layer in the eastern tropical Pacific, parts turn southeast and supply the upwelling off Peru (as suggested by Wyrtki, 1963), while the largest part of this water is integrated into the South Equatorial Current and flows back to the west, or supplies the upwelling at the equator. The Peru Current is much too weak to supply the strong South Equatorial Current, and its high transports can only be accounted for if the Undercurrent provides the necessary water. Detailed observations in the region to the south and west of the Galapagos Islands, which could confirm this process, have still to be made.

## EQUATORIAL COUNTERCURRENT

The Equatorial Countercurrent flows east across the entire Pacific Ocean between the North and the South Equatorial Currents and its existence has been known for a long time. The seasonal variation of this current has been discussed by Knauss (1958). The speed, width, and transport of this current vary considerably with the seasons, as well as over short periods. In recent years several expeditions have made intensive current measurements in sections across the current, which are summarized by Knauss (1961). He finds that the current is very variable, even from day to day, that it is in approximate geostrophic balance, has a sub-surface velocity maximum due to opposing wind stress at the surface, and that appreciable velocities extend to about 150 m depth. When calculating transports, Knauss (1961) used Swallow-float measurements to determine the velocity in layers below 300 m, and for the layer above 1500 m depth found transports of $60 \,.\, 10^{12}$ cm³/sec in August 1958 during Expedition Doldrums (Knauss and Pepin, 1959), and $1 \,.\, 10^{12}$ cm³/sec in July 1959 during Expedition Dorado. Although Knauss was fully aware of the implications of this choice of a reference velocity in greater depths, and though the accuracy of the deep current measurements is not questioned, one must still stress that it is not permissible to calculate



the transport of the Countercurrent above a certain deep reference level; calculations should be restricted to the surface layer, since this current is basically a surface current. The geostrophic transports of the Countercurrent in the upper 300 m relative to 1000 decibars calculated for nine sections are listed in Table II. Montgomery and Stroup (1962) even calculate the geostrophic flow for the entire equatorial current system relative to 300 decibars, thus assuming a really shallow current system. The transports in the upper

### TABLE II

*Geostrophic transports of the Equatorial Countercurrent between the surface and 300 m depth relative to 1000 decibars, during nine crossings of the current*

| Expedition | Date | Longitude | Transport ($10^{12}$cm³/sec) |
|---|---|---|---|
| Equapac | August, 1956 | 135°W. | 18·0 |
| Eastropic (POFI) | November, 1955 | 121° | 13·0 |
| Dorado | July, 1959 | 120° | 5·6 |
| Eastropic (BAIRD) | October, 1955 | 116° | 10·3 |
| Eastropic (BAIRD) | October, 1955 | 116° | 8·7 |
| Doldrums | August, 1958 | 106° | 17·3 |
| Eastropic (HORIZON) | December, 1955 | 103° | 15·7 |
| Step-I | December, 1960 | 95° | 6·5 |
| Eastropic (BAIRD) | November, 1955 | 90° | 7·2 |

300 m for the two sections discussed by Knauss are 17·3 and $5{\cdot}6 \,.\, 10^{12}$ cm³/sec to the east, respectively. This table chiefly demonstrates the strong variation of the transports in this current in space and time, but does not show any systematic decrease of the transports downstream. Four consecutive sections across the Countercurrent in 1951 at 140° W. have shown that in a five-week period the geostrophic transport relative to 400 decibars increased from 18·3 to $22{\cdot}6 \,.\, 10^{12}$ cm³/sec (Austin, Stroup and Rinkel, 1956). This was related to progressive tilting of the thermocline. The maximum speed at the sea surface doubled during this time from 63 to 120 cm/sec.

The transfer of water across the southern and northern boundaries of the Countercurrent seems to be variable and to be dependent on short-term fluctuations of the intensity of the current in time and space. In general there seems, however, to be an intake of water across its southern boundary and a loss of water across its northern boundary. Off the coast of Central America the current splits off, and one branch, usually the stronger, turns north around the Costa Rica Dome forming the Costa Rica Coastal Current which feeds the North Equatorial Current. The flow around the Costa Rica Dome is about $20 \,.\, 10^{12}$ cm³/sec, and the depth of the current increases to about 600 m in the Costa Rica Coastal Current. Only about 40% of this flow takes place in the upper 200 m, while the transport of the Countercurrent is almost completely concentrated in the upper 200 m (Wyrtki, 1964a). No reliable estimate on the percentage of the water of the Countercurrent turning north or south can yet be made.

## NORTH EQUATORIAL CURRENT

The North Equatorial Current is formed by water from the California Current, by water from the Countercurrent, and by water ascending in the eastern tropical Pacific. At 140° W. the North Equatorial Current carries 20–25 . $10^{12}$ $cm^3/sec$ to the west between 10° N. and 25° N. Most of this flow is concentrated in the upper 300 m. The California Current contributes about 12 . $10^{12}$ $cm^3/sec$, the Countercurrent about 8 . $10^{12}$ $cm^3/sec$, the contribution of both currents varying with the season. Computations of the heat and salt balance of the surface and sub-surface water in the eastern tropical Pacific suggest that there is considerable upwelling over the entire region, contributing about 5 . $10^{12}$ $cm^3/sec$ to the surface circulation. A certain loss from the region occurs through the sub-surface flow to the north in the California Current.

## SOUTH EQUATORIAL CURRENT

The South Equatorial Current is developed as a result of the southeast trades in the South Pacific Ocean. Since the trade winds extend across the equator into the northern hemisphere the current also extends to the north of the equator. Its northern boundary is the Equatorial Countercurrent in approximately 4° N. In the south it extends as a well defined current to about 10° S. but continues as a general drift with much smaller and more irregular velocities to the west and southwest. The highest velocities, of more than 50 cm/sec, are found near the equator where the flow is slightly divergent. At the equator the South Equatorial Current is also shallowest, only 20 to 50 m thick, the eastward flowing Equatorial Undercurrent being situated underneath. Farther to the south the South Equatorial Current becomes considerably thicker and extends to about 200 m depth.

The transport of the South Equatorial Current between the Countercurrent in the north and 20° S. is about 45–60 . $10^{12}$ $cm^3/sec$. Most sections for which transports have been calculated do not, however, cover the entire current. At 150° W. Montgomery and Stroup (1962) calculate a transport of 63 . $10^{12}$ $cm^3/sec$ between 6° N. and 7° S., of which half flows north of the equator and half south of it. Calculations for a section at 121° W. give a flow of 20 . $10^{12}$ $cm^3/sec$ between the equator and 7° N. and 23 . $10^{12}$ $cm^3/sec$ between the equator and 8° S. Another section at 130° W. gives a transport of 25 . $10^{12}$ $cm^3/sec$ between the equator and 18° S., and a section at 135° W. gives a transport of 39 . $10^{12}$ $cm^3/sec$ between the equator and 19° S. in the upper 400 m.

It is quite obvious that these enormous transports cannot be supplied either by the Peru Current or the small amount of water of the Countercurrent turning south off the coast of Central America. Further, the pattern of surface currents in the area south of the Galapagos Islands shows the considerable increase of the speed and width of the westward flow from the Peru Current to the South Equatorial Current. If the Peru Current contributes 14 . $10^{12}$ $cm^3/sec$ and the Equatorial Countercurrent about 6 . $10^{12}$ $cm^3/sec$ to the South Equatorial Current, approximately 30 . $10^{12}$ $cm^3/sec$ are still needed to supply this strong current. Such an amount of water can only come from the Equatorial Undercurrent, which transports about 35 . $10^{12}$ $cm^3/sec$ to the east and discharges this amount of water near the Galapagos Islands. It must be assumed, therefore, that water coming directly or indirectly from the Undercurrent feeds the South Equatorial Current. The processes by which this sub-surface water is converted into the water of the South Equatorial Current have still to be studied. Parts of the water of the Undercurrent seem to come into the Peru Countercurrent and feed the upwelling off northern Peru. Other parts definitely participate in the equatorial upwelling to the west of the Galapagos Islands, but the amount of water upwelling along the equator certainly does not exceed a few $10^{12}$ $cm^3/sec$, otherwise the thermal structure would break down. Some of the sub-surface water of the Undercurrent will continue to remain sub-surface water and will be integrated as such in the South Equatorial Current. It seems likely, however, that the upwelling area off the coast of Peru extends northwest towards the Galapagos Islands and is directly connected with the equatorial upwelling west of these islands, at least during the period May through November.



## TRANSVERSE CIRCULATION IN THE EQUATORIAL CURRENT SYSTEM

The four currents mentioned so far form the equatorial current system, which is subject to a very distinct transverse circulation. The principles of this transverse circulation have been derived by Defant (1936) for the Atlantic Ocean and Sverdrup, Johnson and Fleming (1946) stated that a corresponding circulation exists in the equatorial Pacific Ocean: in both cases this transverse circulation was discussed before the discovery of the Undercurrent, and it is therefore surprising that their interpretation of the circulation still remains basically valid. This transverse circulation consists of a divergence in the surface flow at the equator due to the change of sign of the Coriolis parameter, of a convergence at the southern boundary of the Countercurrent, and of another divergence at the northern boundary of the Countercurrent. Between the divergences and convergences circulation cells are developed which are superimposed on the main zonal flow. In each cell the surface flow is from the divergence to the convergence. The sub-surface flow takes place within the upper portions of the discontinuity layer and is in opposite direction.

Cromwell (1953) has also discussed this transverse circulation and strictly applies the principles of isentropic analysis, disregarding the possibility that near the sea surface water masses can be considerably modified by the climatic conditions, and he arrives at conclusions, that are in part contrary to those of both Defant and Sverdrup. The ascending movements in the intense divergences at the equator are called upwelling by Cromwell, and he states that there is no upwelling along the northern boundary of the Countercurrent. He also concludes that some sinking takes place at the southern boundary of the Countercurrent, as long as the equatorial front is developed, but adds that there is otherwise no evidence of sinking or convergence. He shows how the system of surface convergences and divergences in the vicinity of the equator should develop under different wind conditions. When the trade winds have a strong southerly component, there should be a divergence slightly south and at the equator, and a convergence north of the equator. This circulation suggested by Cromwell (1953) may actually develop near the Galapagos Islands (Fedorov, 1963), where southerly winds are frequent, but will not be the normal case in the central Pacific, because it is not consistent with the transverse circulation required by the Undercurrent discovered shortly afterwards. The circulation pattern given by both Defant and Sverdrup, on the other hand, is compatible with the Undercurrent.

When discussing thermocline topography in the eastern Pacific Ocean, Cromwell (1958) described the horizontal distribution of the various convergences and divergences and the associated vertical movements, and tried to elaborate a distinction between upwelling and ridging. However, this distinction seems to reflect the *intensity* of the divergence rather than the *type* of the associated circulation. Along the equator, surface flow is divergent, ascending movements (upwelling) take place, and the flow in the discontinuity layer is convergent. This symmetry may be disturbed temporarily by transient changes in the wind pattern. The upwelling is usually strong enough to cause certain isotherms to intersect the sea surface. Along the southern boundary of the North Equatorial Current the wind-induced surface flow is also divergent, and the thermocline is in a very shallow position, although no intersection of isotherms with the sea surface takes place. This situation is called ridging by Cromwell (1958).

There is some uncertainty about the transverse circulation within the Countercurrent which, as stated above, is rather variable. Because the Countercurrent coincides with the belt of variable and weak winds, it cannot be expected to have a very pronounced wind-induced transverse circulation, as is the case with both the North and South Equatorial Currents. There is even some evidence from surface current charts (Wyrtki, 1965b) and from the dynamic topography of the sea surface (Reid, 1961) that the Countercurrent gains water across its southern boundary and loses water to the north. This would mean that the transverse circulation in the Countercurrent, suggested by Defant (1936) and Sverdrup, Johnson and Fleming (1946) to be a cellular circulation with surface flow to the equator, is weak and probably changeable and consequently the divergence at its northern boundary will also be weak (as also stated by Austin, 1960) and the convergence at its southern boundary may be non-existent during certain times. The strength of the cellular circulation in the Countercurrent is over-emphasized by both Defant and Sverdrup, and this is what Cromwell (1953, 1958) probably wanted to emphasize.

The ridge in the topography of the thermocline along the northern flank of the Countercurrent extends almost to the coast of Costa Rica; off this coast a cyclonic circulation is developed, in which parts of the Countercurrent turn into the North Equatorial Current. The hydrographic structure is that of a thermal dome, as described by Cromwell (1958), who explains the enrichment of the surface layer by mixing, but recognizes the necessity of ascending movements for the maintenance of the dome. In 1959 Expedition Costa Rica Dome was undertaken to study the dome in detail, and Wyrtki (1964a) has explained the upwelling in the centre of the dome by the transverse circulation connected with the cyclonic flow around the dome. The dome appears to be in thermal balance, upwelling adding $7 \cdot 10^{10}$ cm$^3$/sec to the surface circulation which transports about $20 \cdot 10^{12}$ cm$^3$/sec around the dome. The amount of phosphate added to the surface layer by the ascending motion agrees with the observed productivity of the area.



## PERU CURRENT

The flow to the north in the eastern parts of the South Pacific anticyclone off the coast of South America is called the Peru Current, or better, the Peru Current System, because it consists of several more or less independent branches, which interact in a rather complicated way. Schott (1931) has discussed this current on the basis of surface observations of currents, temperature, and salinity. The results of a survey of the coastal waters off South America by the WILLIAM SCORESBY in 1931 are discussed in detail by Gunther (1936), who distinguishes between a Peru Coastal Current, and a Peru Oceanic Current farther offshore, finds a sub-surface countercurrent carrying water of higher salinity south, and states that upwelling is a shallow process, bringing water from an average depth of only 130 m to the surface, and that its structure is spotty, and its appearance irregular in time and space. Sverdrup, Johnson and Fleming (1946) add to Gunther's findings that the northward transport of the Peru Current is $10$–$15 \cdot 10^{12}$ cm$^3$/sec, and that the water of the sub-surface countercurrent is probably of equatorial origin. Neither Gunther nor Sverdrup object to the opinion of Schott (1931) that the Peru Coastal Current converges with the Countercurrent close to the equator. This cannot, however, be true because the Peru Coastal Current leaves the coast near 5° S., its northern boundary is marked by the strong equatorial front, and the east-flowing Countercurrent is always situated north of 4° N. Between these two currents tropical surface water, although originating chiefly from the Countercurrent, flows mainly west.

In 1960, the Step-I Expedition made a comprehensive survey of the entire region off Peru. Wooster and Gilmartin (1961) were able to confirm the sub-surface flow to the south by direct current measurements and by geostrophic calculations. Based on the data of this expedition and on surface wind stress, Wyrtki (1963) has calculated the horizontal and vertical field of motion in the Peru Current. Situated between the Peru Coastal Current and Peru Oceanic Current is the Peru Countercurrent flowing south chiefly as a sub-surface current. At the surface it is usually concealed by the wind drift to the northwest and west, and therefore surface flow to the south is only occasionally observed and seems to be irregular. The Peru Coastal Current transports about $6 \cdot 10^{12}$ cm$^3$/sec to the north across 24° S. and reaches to about 15° S., where most of its water has turned away from the coast. North of 15° S. the wind drift is still to the northwest, but it is shallow, and the southerly flow of the Peru Undercurrent is found immediately beneath the shallow surface layer.

The Peru Countercurrent flows almost due south along 80° W. It is strongest near 100 m depth, but reaches to about 500 m. At 5° S. it transports about $10 \cdot 10^{12}$ cm$^3$/sec, but its transport decreases rapidly to $6 \cdot 10^{12}$ cm$^3$/sec at 15° S. and to $2 \cdot 10^{12}$ cm$^3$/sec at 22° S. This current carries equatorial subsurface water of high salinity and low oxygen content and supplies most of the water engaged in the upwelling processes along the coast north of 15° S. The system of the Peru Coastal Current, the wind drift to the west and the sub-surface Peru Countercurrent are the links in the circulation which maintain the upwelling along the coast. South of 15° S. the upwelling is supplied by the water in the lower layers of the Peru Coastal Current, which is of relatively low salinity. North of 15° S. the upwelling is supplied by water of higher salinity flowing south with the Peru Countercurrent. Both these water masses ascend to the surface along the coast. The total divergence of the wind drift off the coast of Peru between 5° S. and 24° S. has been estimated to be $4 \cdot 6 \,.\, 10^{12}$ $cm^3$/sec which is partly compensated by convergent horizontal flow and partly by upwelling. Along the coast, upwelling is a very shallow process and ascending movements are restricted to the upper 100 m. Farther offshore, ascending movements reach to much deeper levels. The maximum of the velocity of the ascending motion is found in the upper parts of the discontinuity layer. The ascending motion in these layers is connected with onshore movements of the water, and the upwelling along the coast is not a local feature but is related to the processes in the entire current system. The area with ascending movements within the range of the discontinuity layer reaches as far as 700 km offshore. The average velocity of the ascending movements in this area is $20 \,.\, 10^{-5}$ cm/sec, equivalent to 5 m/month; there is no doubt, however, that upwelling along the coast and particularly under the action of strong winds will be much greater during certain short periods. The total contribution of upwelling in the area off Peru is approximately $3 \,.\, 10^{12}$ $cm^3$/sec. Most of this water is supplied in sub-surface movements from the north and north-west.

The Peru Oceanic Current seems to have little interaction with the complicated processes closer to the coast. At 24° S. and to the west of 82° W. it flows northwards and carries about $8 \,.\, 10^{12}$ $cm^3$/sec. North of this position it starts to turn northwest and west, is intensified by water from the Peru Countercurrent, and leaves the area transporting about $14 \,.\, 10^{12}$ $cm^3$/sec. In contrast to the much shallower processes closer to the coast, the Peru Oceanic Current extends with appreciable velocities and transports to about 700 m depth.

Six different areas of upwelling along the coast of Peru are discussed by Schweigger (1958), in contrast to the four areas claimed by Schott (1931) and Gunther (1936), but it seems that such a classification emphasizes transient conditions, concentration of observations, and the natural variability of the upwelling rather than real regional differences along the coast. There is no doubt that the configuration of the coast and of the narrow shelf will cause anomalous local conditions, but the disagreement of the three authors reflects the fact that the selection of these areas is very subjective. Wooster (1961b) documents yearly changes in the temperature of the surface water along the coast of Peru, and points to the marked anomalies during years when El Niño occurs. A comparative study of all eastern boundary currents in the oceans and of the connected upwelling has been published by Wooster and Reid (1963). The influence of the Peru Current on climate and landscape in the coastal region of South America has been dealt with by Schweigger (1959) and by Gierloff-Emden (1959) from a geographical point of view.

## EL NIÑO

The entire area off Peru and as far north as the equator has a considerable annual variation of sea surface temperature, namely, 5° to 7° C. This variation is largely caused by local heating, but in its northern parts it is intensified by the southward shift of the warm tropical surface water north of the equatorial front from December to February. During this period, water of less than 23° C is usually found only along a small strip along the coast of Peru south of 5° S., indicating the upwelling. In some abnormal years the warm water reaches much farther south, the strip of cooler water disappears and this situation, known under the name El Niño, has catastrophic consequences on the biological situation in the region and on the climate along the coast. Such events have been recorded and described for 1891 (Schott, 1931), 1925 (Murphy, 1926), 1941 (Lobell, 1942; Schweigger, 1942), 1953 (Wooster and Jennings, 1955; Posner, 1957), and 1957–58 (Wooster, 1960). Since none of these events has been properly documented by oceanographic observations the phenomenon has not yet been conclusively explained, although several attempts have been made. Schott (1931) explains the occurrence of El Niño as a southward shifting of the Countercurrent leading to a flow of tropical surface water into the region off northern Peru. He also points out (Schott, 1951) that an El Niño occurs simultaneously with very strong upwelling in the Gulf of Panama, as indicated by low surface temperatures. He attributes the whole phenomenon to exceptional weakness of the southeast trades and a displacement of the inter-tropical convergence southwards beyond the equator. Posner (1957) suggests that a substantial shift of the atmospheric pressure and consequent wind systems is responsible for the development of El Niño, and Wooster (1960) attributes it to a weakening of the circulation, which would cause a cessation of upwelling, the formation of a thin warm surface layer by local heating with an associated shallow and sharp thermocline, as well as shoreward movements of warmer water. Analysing more comprehensive data, Bjerknes (1961) arrives at the conclusion that the meteorological control over the occurrence of El Niño lies in the fluctuating strength of the trade winds, but that an accumulation of large masses of warm water in the eastern tropical Pacific Ocean during a longer period is an important prerequisite for the occurrence of El Niño. Of local importance is a weakening of the southerly winds off Ecuador. Although these explanations are plausible and reasonable, it will not be possible to confirm them unless systematic atmospheric and oceanic observations can be carried out during the occurrence of an El Niño.



## GENERAL ASPECTS OF THE CIRCULATION

The theory of wind-driven ocean currents in a baroclinic ocean developed by Sverdrup (1947) and Reid (1948) shows that the zonal equatorial circulation is essentially wind-driven, although computed and observed transports hardly show more than qualitative agreement (Knauss, 1963). Since this theory is linear and frictionless, the Equatorial Undercurrent does not appear at all. Roden (1962) computes total integrated mass transports in the eastern part of the equatorial Pacific Ocean, but his maps show only partial relation to the known circulation pattern. The same is true for seasonal maps of total mass transports calculated by Wyrtki (1964b), who discusses this discrepancy and points out that circulation in the surface layer, which is usually strong, and total mass transports cannot be compared without considerations of the vertical structure of the circulation. Within the Peru Current, where this discrepancy is most pronounced, an almost inverse circulation pattern in the surface and deep layers comprises the total mass transports, which are largely opposite to the surface circulation.

The results of the previous sections can be summarized and combined to construct a circulation system for the eastern equatorial Pacific Ocean, which shows the interaction of the various currents and their mass transports. The transports quoted in Figure 10 are certainly subject to seasonal variations, but are thought to represent approximately the conditions during the period June through December. At 140° W., where the zonal circulation characteristic for the central equatorial Pacific Ocean is fully established, the Undercurrent carries about $35 . 10^{12}$ cm³/sec, and the Countercurrent about $15 . 10^{12}$ cm³/sec to the east. This eastward flow is opposed by about $50 . 10^{12}$ cm³/sec flowing west in the South Equatorial Current and $27 . 10^{12}$ cm³/sec in the North Equatorial Current, giving a total transport of about $27 . 10^{12}$ cm³/sec to the west in the equatorial region. This flow is supplied in about equal parts

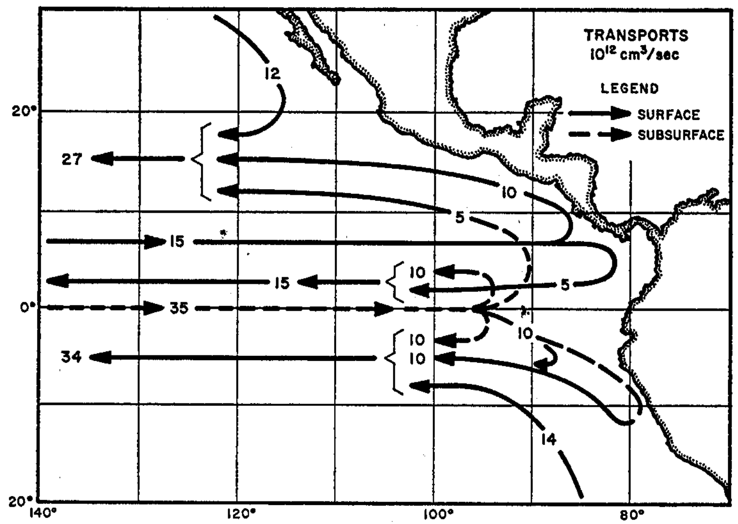


Fig. 10.—Transports in the different branches of the circulation in the eastern equatorial Pacific Ocean during the period from June to December in $10^{12}$ cm³/sec.

by the California and Peru Currents. When the Equatorial Undercurrent approaches the Galapagos Islands and disintegrates, its water spreads north and south. The larger part of the water turning south is integrated into the South Equatorial Current and remains sub-surface water, while the remainder penetrates below the upwelling area off northern Peru and supplies the upwelling. South of the Galapagos Islands all the water of the Undercurrent turning south is integrated into the South Equatorial Current resulting in a tremendous intensification of this current during its passage from the coast of Peru to about 100° W. More than half of the water of those parts of the South Equatorial Current that are situated south of the equator are supplied from the Undercurrent and less than half from the Peru Current.

The Equatorial Countercurrent splits off when approaching the coast of Costa Rica; the larger part turns north into the North Equatorial Current, a smaller part turns south. The surface water turning south from the Countercurrent and the sub-surface water turning north from the Undercurrent



jointly form the water of those parts of the South Equatorial Current situated north of the equator. Approximately $5 . 10^{12}$ cm³/sec of the water of the Undercurrent, which turns north, penetrates into the eastern tropical Pacific as a salinity maximum. There it is engaged into upwelling processes and is integrated into the North Equatorial Current. This current is supplied from the California Current, from the Countercurrent, and partly from the upwelling of water in the eastern tropical Pacific.

Summarizing, it can be said that the wind-driven westward flow of about $75 . 10^{12}$ cm³/sec is opposed by an eastward flow of about $50 . 10^{12}$ cm³/sec, which is primarily caused by the zonal pressure gradient, and the difference in transport is made up by meridional flow towards the equator in the eastern boundary currents.

## STUDIES OF RESTRICTED AREAS

The Askoy Expedition in 1941 investigated the Panama Bight. Wooster (1959) has described the hydrographic structure of the Bight, discussing the changing pattern of surface temperature and salinity and pointing to the occurrence of upwelling. Schaefer, Bishop and Howard (1958) investigated the relations between surface temperature, sea level, and wind, in the Gulf of Panama and demonstrated a high correlation between these factors; upwelling in the Gulf from January to April is caused by strong northerly winds and accompanied by low sea level. Fleming (1940) has inferred that during the upwelling period about 75 m of surface water are driven offshore and replaced by cooler sub-surface water. The effects of the upwelling on the distribution of biota in the Gulf have been studied by Forsberg (1963). A statistical analysis of sea level variations in Panama is given by Roden (1963).

In 1956 and 1957 seven cruises were made into the Gulf of California, and the results were analysed by Roden and Groves (1959). Earlier work was reported by Sverdrup (1941) and summarized by Roden (1958, 1964). Seasonal variations in the upper layers are very large in the northern and central parts of the Gulf. Evaporation over the Gulf is large and totals $5 . 10^9$ cm³/sec. Surface water of increased salinity leaves the Gulf in the surface layer, while deeper water of lower salinity enters the Gulf. The exchange of water between the Gulf and the open Pacific Ocean has been estimated by Roden and Groves (1959) as about $1 . 10^{12}$ cm³/sec in either layer. Upwelling seems to occur in several locations in the Gulf according to the wind conditions.

The oceanographical and biological conditions in the Gulf of Tehuantepec have been observed during four cruises in 1958 and 1959 and the conditions are described by Holmes and Blackburn (1960) and Blackburn (1962a). During the period from September through May strong northerly winds blowing over the isthmus of Tehuantepec cause offshore flow and a doming of the discontinuity layer. The strong winds tend to deepen the mixed layer, especially above the dome, and lead to an enrichment of the surface layer with nutrients. The consequences on the biota and the productivity of the area have been studied, and Blackburn (1962a) finds a close relation between the seasonal variation of the winds and biological activity. However, the centre of highest zooplankton crop is displaced downstream from the centre of highest productivity and strongest upwelling. Roden (1961) has discussed the circulation and the related upwelling in the Gulf from a theoretical point of view, and finds that the wind-induced divergence of the horizontal mass transport would cause an ascending motion of 10 m/day at the bottom of a 100 m thick surface layer. The response time of the system to variable wind conditions should be 1 to 2 days. The surface layer is, however, only 30 m thick, and consequently the ascending movements should not exceed 3 m/day. The ascending motion quoted by Roden (1961) would cause an outcropping of the thermocline after three days of strong winds and such has not been observed.

In order to investigate possible reasons for the accumulation of tuna in the vicinity of oceanic islands and banks, Bennett and Schaefer (1960) have studied the oceanographic situation in their surroundings. The data showed the development of eddies in the vicinity of these islands, a spreading of the thermocline, and possibly more intense mixing. Although some of these features may influence the behaviour of tuna, no conclusive results could be derived.

### CIRCULATION AND FERTILITY

Relations between the topography of the thermocline, the circulation, and the zooplankton crop have been discussed by Brandhorst (1958b). Enrichment of the surface layer is in general associated with a thin mixed layer, which is characteristic for the equator, the northern boundary of the Countercurrent, the Costa Rica Dome, and several places along the coast of Central America. Maps of circulation, and of phosphate and zooplankton distribution for the entire Pacific Ocean by Reid (1962) demonstrate the very high fertility of the eastern equatorial Pacific Ocean. Relations between primary production, chlorophyll, zooplankton, and features of the circulation and the hydrographic structure have been discussed by Holmes, Schaefer and Shimada (1957) and by Holmes (1958a). Brandhorst (1958a, 1959) has analysed the nitrite distribution in the eastern tropical Pacific Ocean and finds that one maximum of nitrite occurs in the thermocline and a second within the oxygen minimum layer. Relations between the abundance of tuna in the eastern tropical Pacific and oceanographical features, with an outlook on possible ways of prediction, have been discussed by Blackburn (1962b, 1965). Locally, relations between the environment and biological factors have been studied in the Gulf of Nicoya by Peterson (1960), and in Pisco Bay by Sears (1954).

Posner (1957) has discussed enrichment processes in the Peru Current and finds that about $8 \cdot 10^5$ tons of phosphate are brought annually into the surface layer, half of this by upwelling, the other half by vertical eddy diffusion. Most authors discuss the fertilization of the surface layer in this part of the ocean under the aspects of ascending movements or wind-induced mixing, but disregard the intense gradient of almost all nutrients within the thermocline, which must lead to a turbulent upward transport of nutrients, even if the vertical eddy diffusivity in the thermocline is only small. As a consequence of the thin mixed layer and of the shallow and sharp discontinuity layer, and the corresponding high position of nutrient-rich sub-surface water, the eastern equatorial Pacific must be a region of a general high fertility, which is so clearly demonstrated in the charts by Reid (1962).



## OUTLOOK ON FUTURE RESEARCH PROBLEMS

Although the general aspects of the oceanography of this region seem now to be well understood and documented and the work of many authors has been combined into a consistent picture of the circulation and other processes governing the oceanic structure of this region, there are still many hypotheses involved, the validity of which will have to be proved by future investigations. One of the major problems of this region is the whereabouts of the water of the Equatorial Undercurrent west of the Galapagos Islands. It will have to be shown by new and more detailed observations how these water masses are re-circulated into the South and North Equatorial Currents. In connection with this problem it will have to be investigated whether or not there is a temporary connection of the upwelling off Peru with the upwelling along the equator. Another, but not at all unrelated phenomenon, is the occasional occurrence of El Niño during the 'wrong' season, but unless comprehensive observations during such an event can be made there seems to be little hope of solving the problem of its causes in a convincing way; moreover, it seems likely that several different combinations of oceanic conditions may lead to its occurrence. The chemical and biological conditions causing the formation of the huge oxygen minima off the coasts of Central America and Peru are far from being understood qualitatively or quantitatively, although it is clear that circulation in these layers is very weak.

Once the general features of the structure and circulation of this region and their average seasonal variations have been understood, future research should increasingly tend to study the deviations from these regular cycles in order to understand the causes of such anomalies, and with the final aim of working out means to predict such anomalies; such predictions are important for long-range weather prediction as well as for those in fishery work. In contrast to these investigations involving the conditions over large areas and the interaction of ocean and atmosphere, other research will be concentrated on the fine-structure of oceanic processes, such as upwelling, eddy formation, and the development of oceanic fronts, and the dependence of these on local or transient atmospheric conditions.

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